# Dynamical Climatology

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# The Annual Cycle of Sea Level in the Eastern Tropical Pacific

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Abstract

The long period response of sea level on the eastern boundary of the Pacific separates, for linear dynamics, into a remotely forced part mainly due to zonal winds along the equator to the west, and a locally driven part where sea level slopes to balance the alongshore wind. Examination of the annual component of sea level on the eastern boundary shows that the locally forced part dominates whereas the remotely forced part plays a major role at semi-annual and interannual periods.

## 1. Introduction

Annual changes of sea level in the tropical Pacific are generally quite small, with amplitudes less than 10cm in most places. Against this background, amplitudes on the eastern boundary are relatively large, reaching 30cm at Balboa in Panama. Another notable feature of the eastern boundary variations is the $180^{o}$ phase difference between the two hemispheres. This phase difference has been ascribed to heating effects (Wyrtki and Leslie, 1980), as it is prominent away from the equator everywhere in the Pacific. However, the phase change on the eastern boundary is particularly distinctive and occurs within a degree or so of the equator. Taken together with the large amplitude on the boundary, other mechanisms seem worth investigating.

The strong annual signal on the eastern boundary has been related to a seasonal Kelvin wave generated further to the west (Kindle, 1979). Busalacchi and O'Brien (1980) used a reduced gravity model to study the seasonal cycle and found a strong semi-annual oscillation at the eastern boundary which correlated well with observed changes in the depth of the $14^{o}$C isotherm. They did not, however, compare their results with sea level changes.

In this paper it will be shown that the long-period sea level response on the eastern boundary of the Pacific ocean can be divided into a remotely forced part and a locally driven part. The latter is due to the alongshore component of the wind which is largely meridional. This

meridional wind component is often ignored in models. However, it is found that the locally driven part is the main contributor to the annual period changes on the eastern boundary, and this explains the larger amplitude and the phase change described above.

The average wind stress in the tropical Pacific for the period June 1979 - May 1982 is shown in Figure 1. The large alongshore component on the eastern boundary is evident, as is the change in sign and the large gradient of this component associated with the Intertropical Convergence Zone. There is also considerable temporal variability in this field associated, among other things, with seasonal and interannual changes in the position of the ITCZ (Wyrtki and Meyers, 1975). Figure 2 shows sea level changes at three near equatorial stations for the period June 1979 to March 1983. The 1982/83 El Nino event distorts the last section of the record, but prior to that annual signals at each station can be seen, and these vary considerably between stations. This is a signature of locally forced changes. Remotely forced changes would be expected to be similar at all three stations.

A simple theory demonstrating the effect of purely meridional wind stress forcing on sea level will first be given and then this will be applied to the observed wind field. The fundamental structure of the observed annual cycle is reproduced well, both in phase and amplitude, including such peculiarities as the signal at Buenaventura. Differences between the theory and observations are related to the semi-annual signal in the pycnocline (Busalacchi and O'Brien, 1980). It seems likely, therefore, that the meridional wind stress balance contributes the major

part of the annual cycle. Remotely forced effects are, however, important for interannual changes. This can be seen from comparing the annual mean response with the response seen in particular years, including the strong 1982/83 event.

## 2. Theory of annual fluctuations on the eastern boundary

The eastern boundary plays a unique role in the behaviour of annual period waves because of the special nature of equatorially trapped long waves. For any given vertical mode, there is only one wave, the Kelvin wave, which can bring information to the boundary from the west. The remaining infinite set of planetary waves can only carry information away from the eastern boundary. Thus the eastern boundary response can be subdivided into two independent constituents: (a) the remotely forced part, i.e. that which is the result of Kelvin waves incident on the boundary, and (b) the locally forced part.

The remotely forced component has received some attention in the past, but the locally forced response has not. This part is driven by the alongshore wind stress and satisfies a very simple equation. That is because coastal Kelvin waves carry information out of the equatorial zone in a very short time (a week or two) compared with the forcing period. Hence the alongshore momentum balance is effectively an equilibrium one. Treating the boundary as being purely meridional, with y as distance northwards of the equator, and u as the normal component of velocity, the equilibrium balance is

$$\beta y u = -\frac{1}{\rho_0}\frac{\partial p}{\partial y} + \frac{1}{\rho_0}\frac{\partial Y}{\partial z} \qquad (2.1)$$

where $\beta$ is the gradient of the Coriolis parameter, p is the pressure perturbation, $\rho_0$ the density of the ocean, Y the horizontal alongshore stress and z the vertical coordinate. If Y is assumed to vary linearly from its surface value $Y_s$ to zero over an upper mixed layer of depth $H_{mix}$ (see e.g. Gill, 1982), then the balance on the boundary (where u=0) at the surface (where $p=\rho_0 g \eta$ ) is

$$g\frac{\partial \eta}{\partial y} = Y_s / H_{mix} \qquad (2.2)$$

which gives the surface elevation apart from an arbitrary constant. It is the determination of this constant that provides the special interest in this problem.

To see how the constant is determined requires a look at the theory of equatorially trapped modes which is given in chapter 11 of Gill (1982). Firstly, the terms in equations such as (2.1) are expanded in terms of vertical modes. For the mth mode, (2.1) gives

$$g\frac{\partial \eta^{(m)}}{\partial y} = \sigma_m Y_s / \rho_0 H_{mix} \qquad (2.3)$$

where $\eta^{(m)}$ is the projection of $p/\rho_0$ onto the mth mode and $\sigma_m$ is the projection of a step function (unity in the mixed layer, zero below) onto the same mode. Secondly, $\eta^{(m)}$ and the other dependent variable projections are expanded in terms of parabolic cylinder functions of order n to give

equations for variables $q_m^{(m)}$ and $r_n^{(m)}$ , which represent the amplitudes of equatorially trapped waves. In particular, $q_o^{(m)}$ represents the Kelvin wave (Gill, 1982). The equation for $q_o^{(m)}$ is obtained without reference to (2.1) and so is not forced by the meridional wind. Therefore it does not contribute at all to the solution of (2.3). At the eastern boundary, $\eta^{(m)}$ is directly proportional to $q^{(m)}$

$$q^{(m)} = \left(\frac{g}{H_m}\right)^{\frac{1}{2}} \eta^{(m)} \qquad (2.4)$$

where $H_m = c_m^2 / g$ is the equivalent depth for mode m and $c_m$ is the corresponding wave speed. So $\eta^{(m)}$ cannot contain a contribution from the zero order cylinder function. This is the condition which determines the arbitrary constant, namely that the projection of $\eta^{(m)}$ onto the zero order cylinder function belongs to the remotely forced contribution rather than the locally forced part.

An alternative way of seeing the arbitrariness which comes when trying to integrate (2.3) is to expand both $\eta^{(m)}$ and $Y_s$ in cylinder functions, as in Gill (1982),

$$g\eta^{(m)}, Y_s/\rho_0 H_{mix} = \sum_{n=0}^{\infty} (\eta_n^{(m)}, Y_n)\, D_n((2\beta/c_m)^{\frac{1}{2}} y). \qquad (2.5)$$

Substitution in (2.3) and equating coefficients gives

$$(2\beta/c_m)^{\frac{1}{2}} \eta_1^{(m)} = \sigma_m Y_0$$

$$(2\beta/c_m)^{\frac{1}{2}} (-\eta_0^{(m)} + \eta_2^{(m)}) = \sigma_m Y_1 \qquad (2.6)$$

etc. The arbitrariness comes in the second equation. The appropriate condition is to choose $\eta_0^{(m)}$ as zero, so the second equation gives a value for $\eta_2^{(m)}$. Hence all the coefficients $\eta_n^{(m)}$ are assigned unique values. It can be shown that if $\eta_0^{(m)}$ were given a non-zero value, it would result in a constant being added to $\eta^{(m)}$.

Hence a precise procedure for calculating $\eta$ is established. The equation (2.3) is integrated for each vertical mode subject to the constraint

$$\int_{-\infty}^{\infty} \eta^{(m)} D_0\left((2\beta/c_m)^{\frac{1}{2}} y\right) dy = 0 \qquad (2.7)$$

and $\eta$, the surface value, is obtained by adding modes. This determines the arbitrary constant involved in integrating (2.2).

As examples of solutions to (2.2) consider

$$Y_s = 2\beta y / K \qquad (2.8a)$$

$$Y_s = 0.01\left(1 - \mathrm{erf}\left((12\beta/1.55)^{\frac{1}{2}} y/5\right)\right). \qquad (2.8b)$$

Then the solution to the forcing of (2.8a) is

$$\eta = \left[\beta y^2/K - \frac{1}{2K}\sum_{m=1}^{\infty} \sigma_m a_m\right] / g\rho_0 H_{mix}, \qquad (2.9)$$

where $a_m$ is the constant determined by equation (2.7), demonstrating the

effect of an equatorially asymmetric wind stress, while (2.8b) gives, assuming that $\sigma_1 = 1$, $\sigma_m = 0$, $m > 1$,

$$\eta' = \frac{0.01}{\rho g H_{mix}} \left[ y - y \operatorname{erf}\left( (2\beta/2.55)^{\frac{1}{2}} y/5 \right) + \frac{5\left(1 - e^{-2\beta y^2/(25.2.55)}\right)}{(2\beta\pi/2.55)^{\frac{1}{2}}} \right.$$

$$\left. + \frac{\left(2(2\beta/2.55)^{\frac{1}{2}} + 5\right)}{2\left[(2\beta\pi/2.55)^{\frac{1}{2}}\left(\frac{1}{4} + 2\beta/25.2.55\right)\right]^{\frac{1}{2}}} \right] \qquad (2.10)$$

illustrated in Figure 3.

The integration constant for the real wind stress was calculated by integrating between $\pm 15^\circ$ of latitude and using only the first mode cylinder functions. Using the result for the example (2.8a) as a guide, the error so incurred was estimated to be less than 1cm, and so not significant.

## 3. The Annual Sea Level Cycle in the Eastern Pacific

The contribution of meridional wind stress to the eastern equatorial Pacific annual sea level signal will now be examined in two ways. Firstly, long term monthly means will be compared with the result of the simple theory of section 2 and, secondly, particular years will be scrutinized. In carrying out these two tests both the general and specific contributions of the meridional wind stress to the annual cycle will be studied. Also, the effect of remote forcing in the region can be assessed

from differences between our theory and observations. The eastern Pacific between the Galapagos Islands and Ecuador will be considered to have a mixed layer of 7-10 metres depth (Levitus, 1982) and a first mode Kelvin wave speed of $2.55ms^{-1}$ . The wind stress field used for the calculations is that of Florida State University. Details of its production are given in O'Brien and Goldenberg (1982) and Inoue and O'Brien (1984). To compute the actual stress values from the tabulated psuedo-stress a density of $1.2kgm^{-3}$ and a drag coefficient of $1.7x10^{-3}$ (Inoue and O'Brien, 1984) were used. The necessary integrals were calculated between 15° S and 15° N by which stage the Gaussian function in the integrals is very small. Sea level data were taken from Wyrtki and Leslie (1980), the Service for Mean Sea Level (1977) and data kindly supplied by K. Wyrtki.

In Figure 4 the observed monthly average sea levels for 5 stations are compared with those given by theory. The first diagram shows the results at the Galapagos Islands. The assumptions of uniform meridional wind stress, with respect to longitude, and the shallow mixed layer depth are not so accurate here, hence we see only a rough fit. However, both the amplitude and basic phase of the sea level variation are predicted. The second diagram, for 4 stations along the South American coast, shows excellent agreement for both phase and amplitude. Even such an unusual signal as that at Buenaventura is well described by meridional wind stress arguments alone. From examining the monthly mean wind stress maps in Wyrtki and Meyers (1975) it is seen now that the large dip in the sea level just north of the equator in the northern spring is due to the rapid southward retreat of the ITCZ. The accompanying relaxation of the winds south of the equator allows a higher sea level in this region, which is

eroded as the wind front moves up the coast again in late spring, piling up water north of the equator.

There is a tendency in these plots for the observed sea level to be lower in the spring than predicted, higher in the summer and lower once more in the fall. Comparison with the pycnocline anomaly diagram of Busalacchi and O'Brien (1980, Fig. 7), which shows a definite semi-annual signal, suggests that the reason for this may be found in the influence on the sea level of disturbances propagating from the west. However, this remote forcing would appear to be only a secondary component of the annual cycle.

It should be noted that for this comparison the periods over which the sea level and wind fields were averaged are different. For the wind field the means were calculated over 1961-1983, while the sea level averages were for various periods, dependent on the lifetime of the station, although in all cases the averaging periods included most of the 1960s. However, as the results of data studies by Barnett (1977), Goldenberg and O'Brien (1981), and Enfield (1981a&b), show little interannual variability of winds in this region the different averaging periods should not cause much aliasing.

Having seen how well the average fields agree we now turn to an examination of particular years. In Figure 5 the observed and predicted sea levels at the 5 stations considered in Figure 4 are shown for the years 1963 and 1964. It is immediately clear that the comparison is considerably worse. In any individual year other events are occurring to

modify the average cycle and, as at San Cristobal, these may be of much larger magnitude than the mean. Along the South American coast, however, the basic shape of the observed variation is reproduced in the theoretical predictions. Obviously there is a lot more going on but magnitudes and phases are roughly right. The divergence between model and observations follows a similar, but amplified, pattern to those in Figure 4. From our earlier discussion this implies that in these years there were stronger than usual signals being fed into the area from the west. This particular period has been noted in the literature as one of considerable activity. In particular, 1963 saw unusually heavy rainfall in the central Pacific (Quinn, 1974) and is classified as a minor El Nino (Enfield and Allen, 1980). However, it has been seen that the meridional winds along the South American coast do not have a large interannual variability and indeed Wyrtki (1975) showed that even during El Nino disturbances the Peruvian coastal winds do not collapse, as was once thought.

To further emphasise the local forcing of the average annual sea level cycle, as opposed to the remote forcing of particular events, the contribution of the meridional wind balance to sea level during 1982 and the first quarter of 1983 is shown in Figure 6. This period was chosen to correspond to that studied by Busalacchi and Cane (1985), using the model of Cane and Patton (1984). They give plots of sea level at a number of places in the equatorial Pacific, both observed and as calculated by their zonal wind-driven, quadruple vertical mode model. Two locations, Santa Cruz on the Galapagos Islands and La Libertad in mainland Ecuador, are in our area of interest. As can be seen the meridional wind contribution to the sea level is completely swamped by the effect of zonal remote forcing.

well modelled by Busalacchi and Cane (1985).

## 4. Conclusions

Using only monthly mean meridional wind stress most of the average annual signal in the eastern tropical Pacific sea level can be explained. This involves a simple theory balancing the sea level and local wind stress forcing, without needing to consider the effect of remote forcing. The significant features, namely the phase and amplitude, on both sides of the equator, are well described. Small differences between theory and observation correlate with the effects of remote forcing from the west.

For individual years, however, and especially those associated with El Nino disturbances, the predictions of the local theory can be quite poor, whereas calculations based on remote forcing are good. Thus interannual variability in sea level is dominated by remote rather than local forcing.

## Acknowledgements

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Figure Legends

Figure 1: Average Pacific pseudo wind stress for June 1979 - May 1982. Very small stresses not shown.

Figure 2: Sea level anomalies in the eastern equatorial Pacific during the period June 1979 - December 1983.

Figure 3: Alongshore sea level for the wind stress of eq. (2.8b). $H_{mix}$ =10m.

Figure 4: Annual average sea level anomalies in the eastern Pacific, both from theory (———) and observations (- - -). $H_{mix}$ =7.5m, c =2.55ms$^{-1}$ .

a) At San Cristobal (Galapagos Islands, 0°54'S, 89°37'W);

b) Along the South American coast at Buenaventura (3°54'N, 77°06'W), Tumaco (1°50'N, 78°44'W), Talara (4°37'S, 81°17'W) and Chimbote (9°05'S, 78°38'W).

Figure 5: Sea level anomalies for the years 1963 and 1964 at the stations used in Fig. 4. For the legend see Fig. 4.

Figure 6: Sea level anomalies at a) San Cristobal (Galapagos Islands, 0°27'S, 90 17'W) and b) La Libertad (2°12'S, 80°55'W) for the period January 1982 - March 1983. For the legend see Fig. 4.

Cf. Busalacchi and Cane (1985, Fig. 4).

AVERAGE PACIFIC WIND STRESS JUNE 1979 - MAY 1982

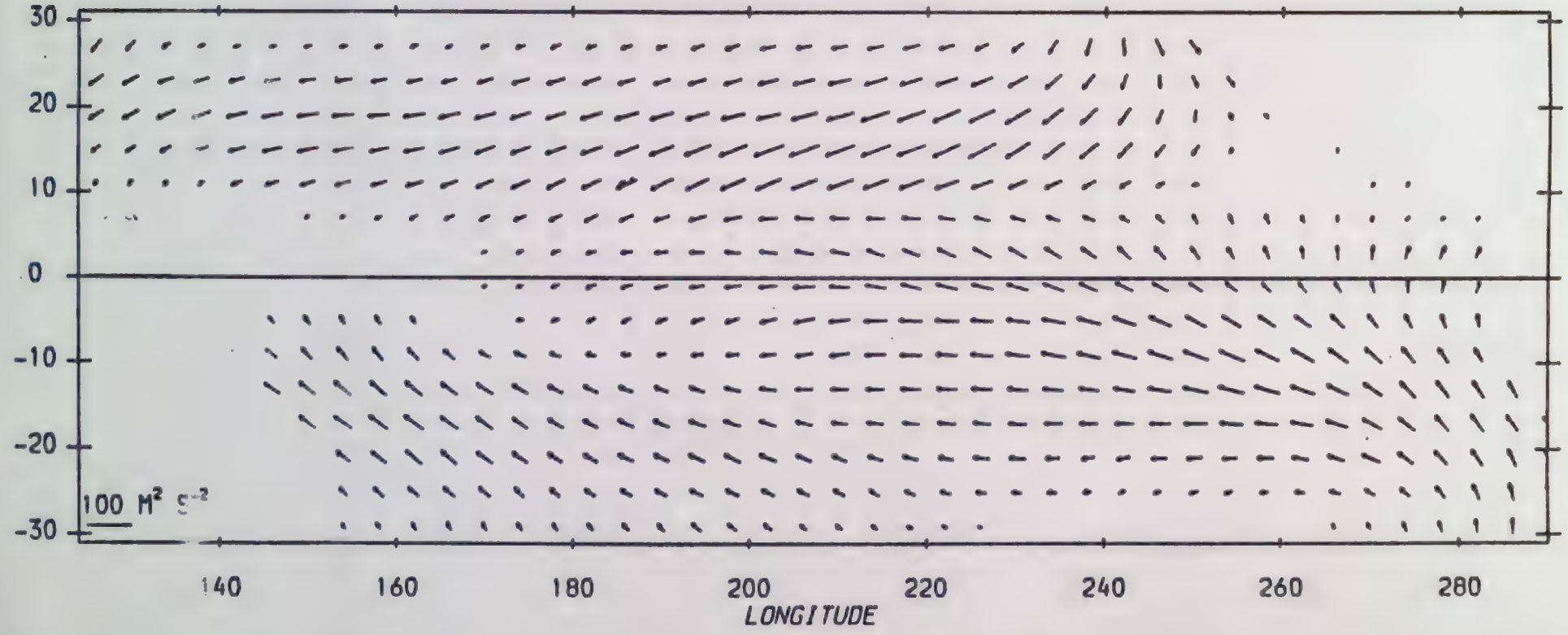



Fig. 1

Fig. 2

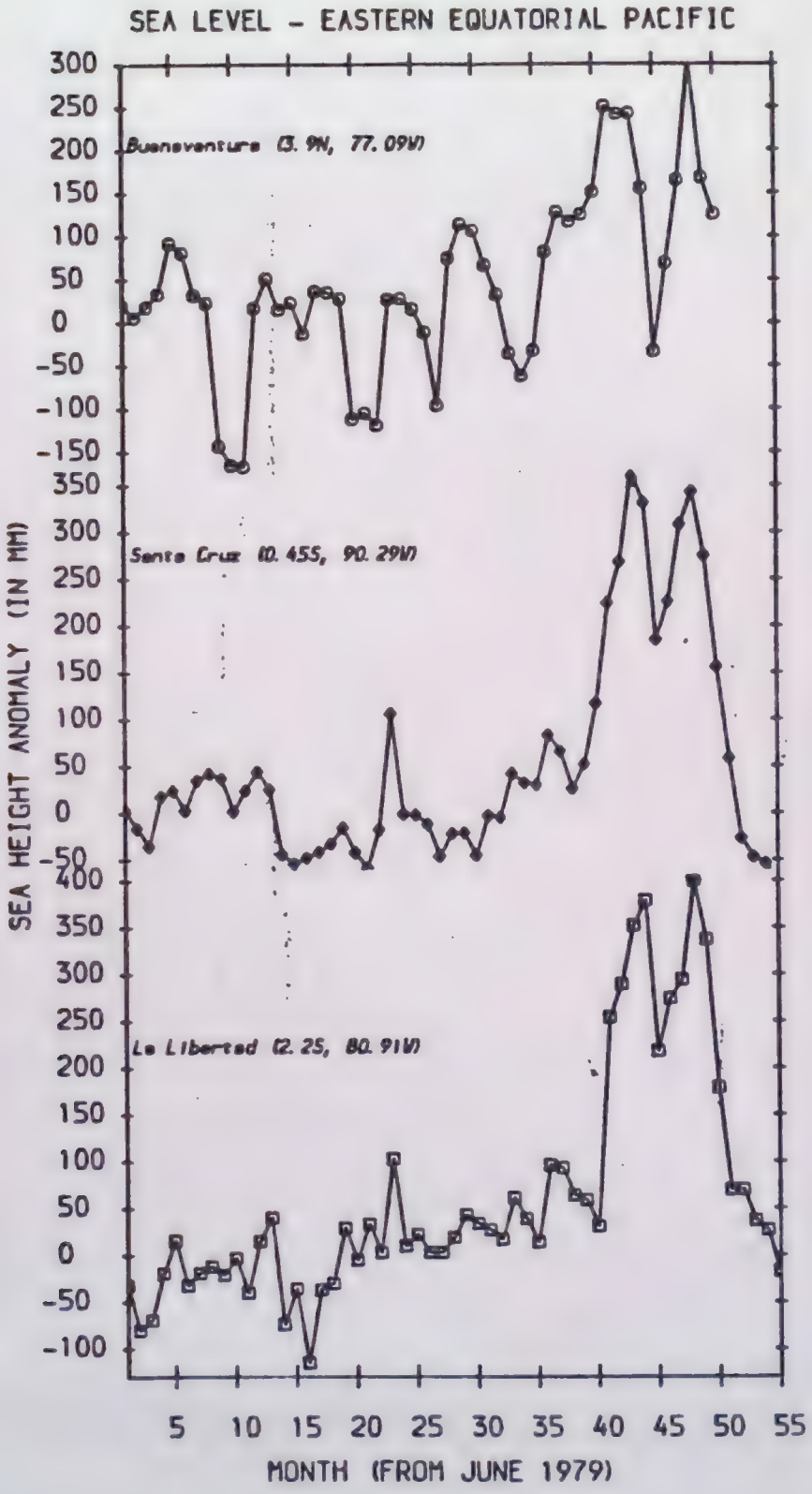
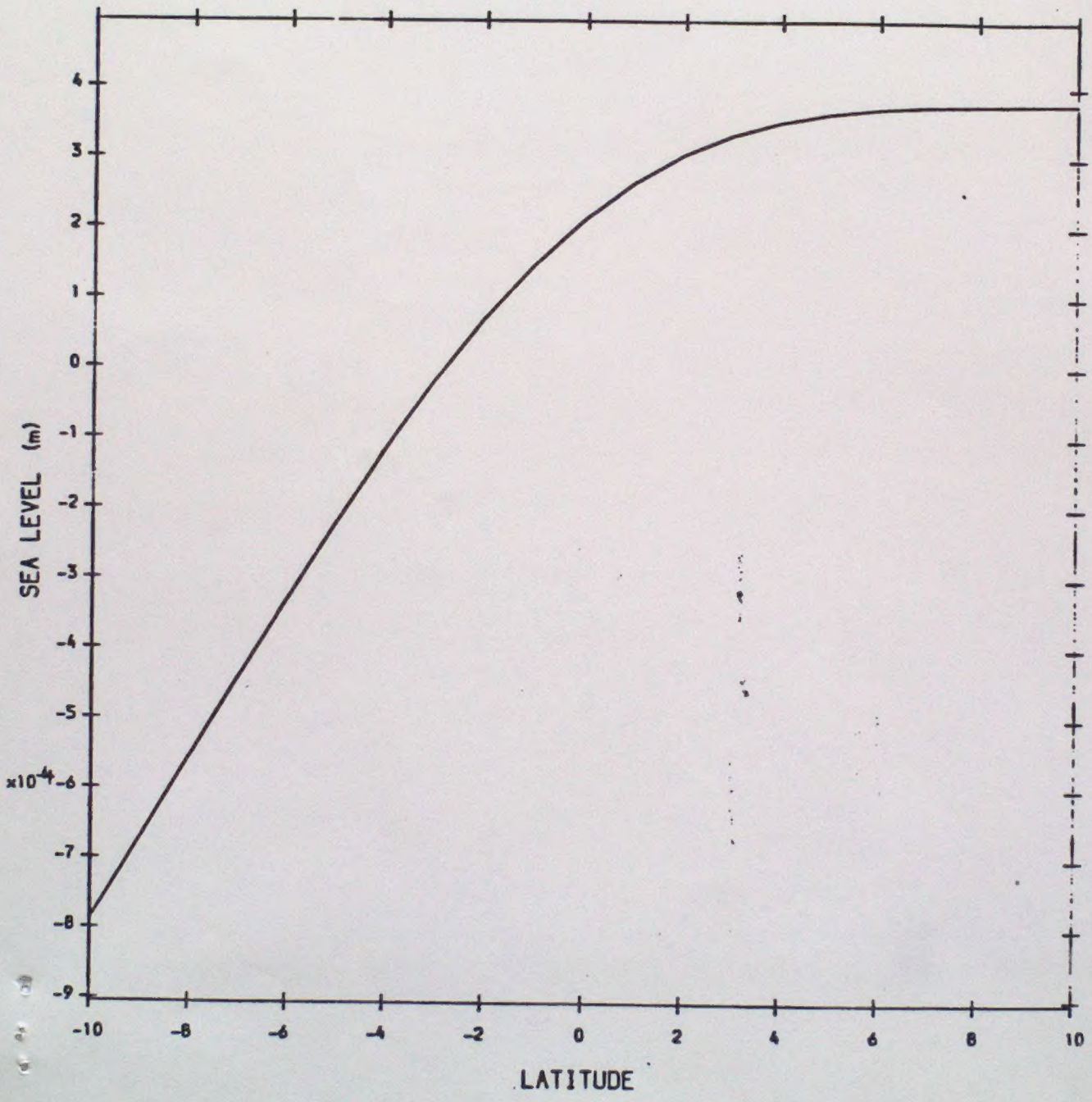
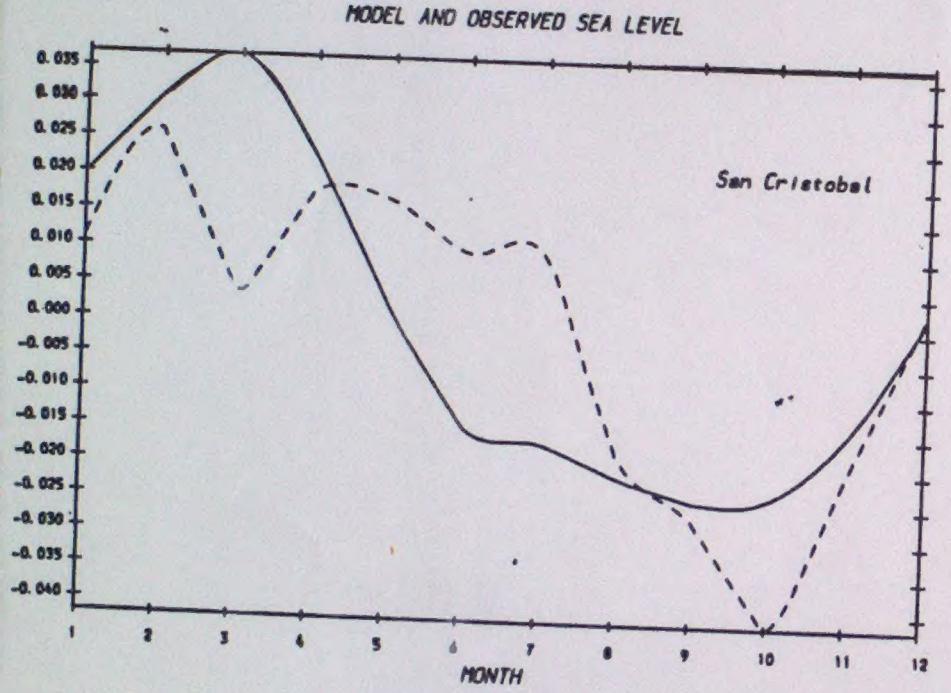
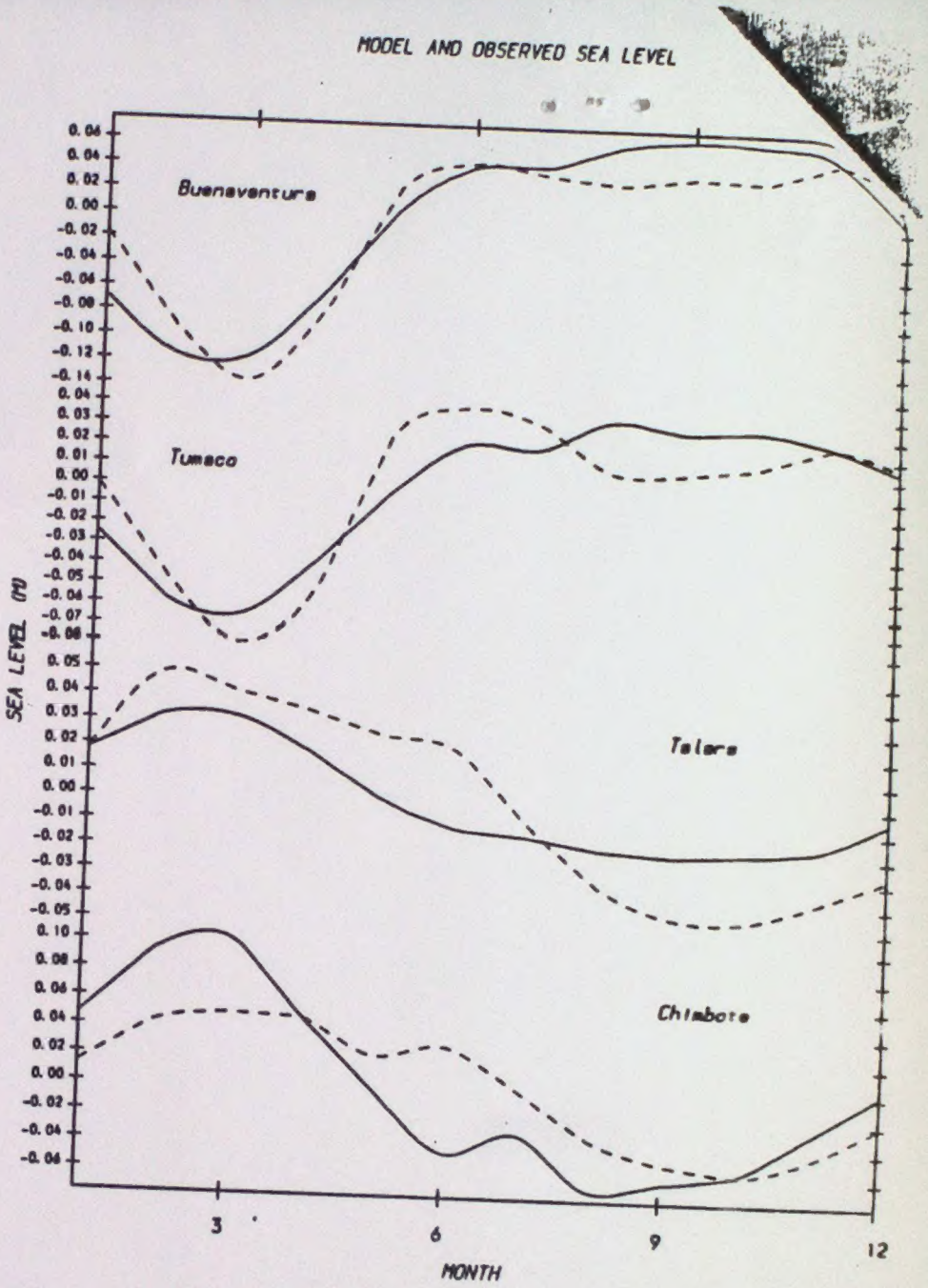

SEA LEVEL - EASTERN EQUATORIAL PACIFIC
Buenaventura (3.9N, 77.09W)
Santa Cruz (0.455, 90.29W)
La Libertad (2.25, 80.91W)
SEA HEIGHT ANOMALY (IN MM)
MONTH (FROM JUNE 1979)

Figure 3

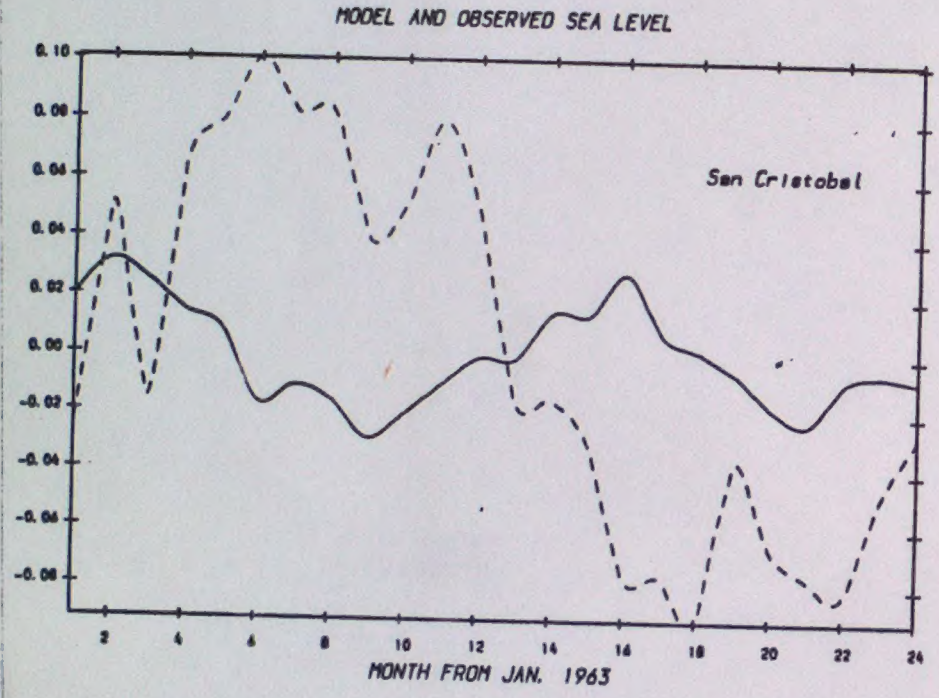



MODEL AND OBSERVED SEA LEVEL
San Cristobal
MONTH

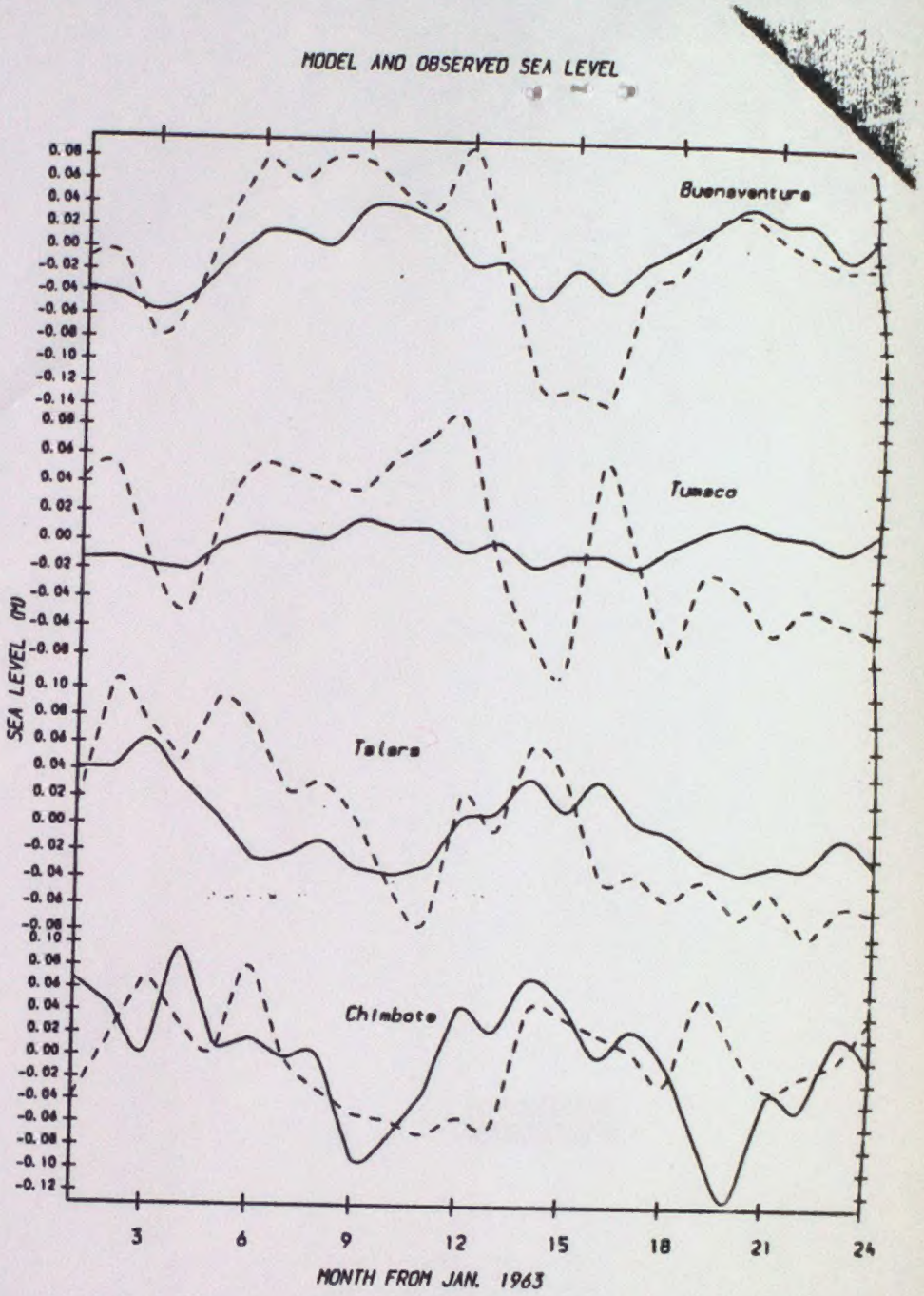
a)


MODEL AND OBSERVED SEA LEVEL
Buenaventura
Tumaco
Talara
Chimbote
SEA LEVEL (M)
MONTH

b)


MODEL AND OBSERVED SEA LEVEL
San Cristobal
MONTH FROM JAN. 1963

a)


MODEL AND OBSERVED SEA LEVEL
Buenaventura
Tumaco
Talara
Chimbote
SEA LEVEL (M)
MONTH FROM JAN. 1963

b)

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